Exploring the theory of plate tectonics: the role of mantle lithosphere structure

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Abstract: This review of the role of the mantle lithosphere in plate tectonic processes collates a wide range of recent studies from seismology and numerical modelling. A continually growing catalogue of deep geophysical imaging has illuminated the mantle lithosphere, and with it generated new interpretations of how the lithosphere evolves. Here, we present a review of the current ideas about the role of continental mantle lithosphere in plate tectonic processes. Evidence seems to be growing that scarring in continental mantle lithosphere is rather ubiquitous, which implies a reassessment of the widely-held view that it is inheritance of crustal structure only (rather than the lithosphere as a whole) that is most important in the conventional theory of plate tectonics (e.g., the Wilson Cycle). Recent studies have interpreted mantle lithosphere heterogeneities to be pre-existing structures, and as such linked to the Wilson Cycle and inheritance. We consider the current fundamental questions in the role of the mantle lithosphere in causing tectonic deformation, reviewing recent results alongside highlighting the potential of the deep lithosphere in infiltrating every aspect of plate tectonics processes.

The reactivation of features formed through previous collisional or rifting events (i.e., inheritance) is a tenet of plate tectonic theory (e.g., Wilson, 1966). Reactivation events occurring along well-defined, pre-existing features such as faults, shear zones or lithological contacts (Holdsworth et al., 1997) are well understood in that they form in preference to new structures (e.g. Sutton and Watson 1986; Butler et al. 1997 and references therein) during continental lithosphere deformation (e.g., major transcurrent fault systems, orogenic belts, and rifted basins in both intracontinental and continental margin settings (White et al., 1986; Handy, 1989; Tommasi et al., 1994, Holdsworth et al., 1997, 2001; Vauchez et al., 1998; Handy et al., 2001; Thomas, 2006)). Furthermore, the migration of hydrous fluids and magmas in continental regions are often through channelways defined by long-lived inherited structures (e.g. see Kerrich 1986; Hutton 1988; McCaig 1997), adding to the importance of pre-existing features in the continental lithosphere. Although discussion of inheritance in the mantle lithosphere has been conducted (e.g., Holdsworth et al., 2001), most research into this topic has focussed on crustal tectonics rather than any deeper structures (e.g., D'Lemos et al., 1997; Holdsworth, 2004; Thomas, 2006).

Compared to the overlying crust, the evolution of the mantle lithosphere is poorly understood; yet, as the main constituent of the lithosphere, this region is fundamental to controlling the tectonic behaviour of the Earth. Although the crust and the mantle lithosphere differ in their chemical compositions, the mantle lithosphere can be distinguished from the sub-lithosphere through mechanical properties related to flow regime. The rheology of the lithospheric layers governs deformation driven by interior forces (Bürgmann and Dresen, 2008), with elastic, plastic (brittle), or viscous (ductile) properties exhibited (Burov, 2011). This layering of the lithosphere is complex, and often unique to the local environment. However, it is important to understand in the context of plate tectonics.

Evidence is growing that heterogeneities within the mantle lithosphere are ubiquitous (e.g., Rawlinson and Fishwick, 2011; Bastow et al., 2013; Schiffer et al., 2014, 2015, 2016; Schaeffer and Lebedev, 2014; Rasendra et al., 2014; Bao et al., 2014; Kahraman et al., 2015; Hopper and Fischer, 2015; Tauzin et al., 2016; Park and Levin, 2016a, 2016b; Biryol et al., 2016; Boyce et al., 2016; Dave et al., 2016). The first-order principles of what this means for past and future tectonic processes are still not clear. However, there are a number of studies offering theories as to what these structures can mean in terms of the wider Wilson Cycle process. Below, we outline broad descriptions of lithosphere rheology to contextualize the arena of study. In the following sections, we highlight the processes involved in the Wilson Cycle (focussing on inherited structures), followed by a discussion on imaging structures in the mantle lithosphere and the difficulty in unravelling the processes required to generate them, culminating in an analysis of recent numerical models and seismic studies that add to the understanding of the role of the mantle lithosphere in the Wilson Cycle. The main focus of the review is to bring together thoughts on the mantle lithosphere and, to begin, we need to understand how the layer behaves rheologically.

Lithosphere rheology

Layering is present within tectonic plates due to the modifying effects of depth-dependent temperature and pressure on rheology. Through the extrapolation of experimental rock mechanics data, yield-strength envelopes can predict the maximum differential stress supported by rock as a function of depth (Goetze and Evans, 1979). By integrating the plastic and ductile conditions of the material within each layer as a function of temperature and pressure, the flow regime of the lithosphere can be estimated. As a result, yield-strength envelopes offer an insight into the mechanical behaviour of lithospheric plates (Burov, 2011).

Bürgmann and Dresen (2008) outlined three food-based analogies to the strength of continental tectonic plates: jelly sandwich; crème brûlée; and banana split (Figure 1). A 'jelly sandwich' strength profile is characterized by a weak lower crust (jelly) between a strong upper crust and mantle lithosphere (bread), as shown in Figure 1a. Relatively cool temperatures in continental interiors generate a strong upper crust (Rutter and Brodie 2004a,b; Rybacki et al. 2006), governed by Mohr-Coulomb theory to produce frictional plastic deformation. The lower crust transitions to viscous flow as temperature and pressure increase, producing a weak ductile layer (Bürgmann and Dresen, 2008). The strength of the jelly sandwich profile lies in the ultramafic mantle (Hirth and Kohlstedt 2003). A 'crème brûlée' profile describes a lithosphere where the strength resides within the crust (Figure 1b), with high temperatures and/or water content weakening the material strength below the crust (Jackson, 2002). The brittle crust produces a deformation regime which acts as the lid to the crème brûlée profile.

Jelly sandwich and crème brûlée can describe the profile within a continental interior (the third profile – banana split – predominately describes plate boundaries and will be discussed below) and have generated some discussion as to the preferred model to be used in geodynamic analysis. Studies into earthquake distribution suggest that continental mantle lithosphere could behave in a ductile manner, with most of the strength of the lithosphere residing in the upper crust (i.e., a crème brûlée rheology) (Déverchère et al., 2001; Jackson, 2002; Maggi et al., 2000). However, laboratory flow laws indicate that the mantle lithosphere would have a complex layering of brittle and ductile material (e.g., Brace and Kohlstedt, 1980; Sawyer, 1985; Gueydan et al., 2014), with a broad consensus in the literature indicating that the mantle lithosphere would be strong enough to support high stresses. Old stable intraplate lithosphere has been interpreted to not have a crème brûlée rheology as it would not maintain the strength and stability to support a craton over long-timescales (Burov and Watts, 2006; Burov, 2010).

The final model is described as a 'banana split' and refers to the changing strength profile across a plate boundary (Bürgmann and Dresen, 2008). Thermal, fluid, and strain-rate processes can combine at tectonic boundaries to weaken the overall strength of the lithosphere (Figure 1c). Major crustal fault zones are taken into consideration with this strength profile, with zones of weakness being generated throughout the thickness of the lithosphere (Bürgmann and Dresen, 2008). Previous studies on mature fault zones (e.g., the San Andreas) have suggested a frictionally weak crust, with weakened shear zones within the viscous regime (Zoback et al., 1987). There are a number of mechanisms that can produce weakening at plate boundaries, such as grain-size reduction (Bercovici and Ricard, 2014; Krajcinovic, 1996; Skemer et al., 2010; Warren and Hirth, 2006; Linckens et al., 2015), that occur through plate tectonic processes related to the Wilson Cycle.

The Wilson Cycle

In 1966, based on evidence in the fossil record and the dating of vestiges of ancient volcanoes, Wilson (1966) proposed a cycle describing the opening and closing of oceanic basins. This cycle provided a method of amalgamating continental material (into a supercontinent) that would be subsequently dispersed (e.g., into a fragmented configuration like the present-day). Wilson (1966), building on previous studies (e.g., Hess, 1962; Vine and Matthews, 1963; Wilson, 1965), outlined a four-stage "Wilson Cycle" (as it was later named by Dewey and Burke (1974)): the dispersal (or rifting) of a continent; continental drift, seafloor spreading, and the formation of oceanic basins; new subduction initiation and the subsequent closure of oceanic basins through oceanic lithosphere subduction; and continent-continent collision and closure of the oceanic basin (Figure 2).

Over the past 50 years this conventional theory of plate tectonics has been at the forefront of geodynamics. However, many features of lithosphere evolution fall outside the realm of the Wilson Cycle: plate tectonics has progressed beyond plate boundaries as the sole locus of major deformation with the study of intraplate orogenesis (e.g., Sykes, 1972, 1978; Smith and Bruhn, 1984; Sibson, 1992; Ziegler et al., 1995, 1998; Stein and Liu, 2009; Stephenson et al., 2009); mantle lithosphere processes generating lithospheric instabilities (in the form of viscous dripping and delamination) that represent a foundering and recycling of plate material (e.g., Bird, 1979; Houseman et al., 1981, 1997; Gögüs and Pysklywec, 2008; Bajolet et al., 2012; Gögüs et al., 2016) in situ mantle lithosphere inversion of Archean cratonic keels (Percival and Pysklywec, 2007); and the interaction of subduction and large low shear velocity provinces in driving the development of large igneous provinces at the surface (e.g., Ernst et al., 2005; McNamara and Zhong, 2005; Bull et al., 2009; Heron et al., 2015a; Mallard et al., 2016).

Among these, the study of intraplate orogenesis has generated several mechanisms for deformation within a plate interior (Figure 2). These mechanisms include pre-existing lithosphere structures, the presence of fluids, the burial of highly radiogenic material and other temperature anomalies, mantle lithosphere instability, compositional strengthening, and strain rate (e.g., Ziegler, 1987; Ziegler et al., 1995, 1998; Sandiford, 1999; Nielsen and Hansen, 2010; Hansen and Nielsen, 2002; Pysklywec and Beaumont, 2004; Sandiford et al., 2006; Stephenson et al., 2009; Heron and Pysklywec, 2016). If intraplate orogenesis can be influenced by similar mechanisms that generate other (established) plate tectonic processes (such as rifting), then it should be recognized as part of plate tectonic theory (e.g., Figure 2).

Inheritance

Experiments on rock properties find that deformation generates weak zones that, over time, can be dormant (or be reactivated) depending on how the material strength is affected by changes in ambient stresses. A reduction in grain size is a characteristic of this lithospheric damage (Bercovici and Ricard, 2014), which can be abundant at tectonic margins in the form of peridotite mylonites (Warren and Hirth, 2006; Skemer et al., 2010). The lithospheric strength of the banana split model (Figure 1c) could be indicative of this weakness at plate boundaries given the rheological impact of the reduced grain size.

The reactivation of structures within the crustal lithosphere has previously been well documented as being part of Wilson Cycle processes (Holdsworth et al., 2001; Holdsworth, 2004). In terms of rifted continents, brittle structures in the shallow crust inherited from previous tectonic events have been interpreted to define the shape of the margin (Thomas, 2006). Furthermore, crustal inheritance could also play a role in intraplate deformation. Stephenson et al. (2009) identified that thermal structures from previous tectonic events could also play an important role in deformation away from plate boundaries in southeastern Ukraine. The continuation of ancient tectonics to influence deformation, even away from active plate boundaries, is a strong indication of the role of inheritance in all forms of plate tectonics.

In discussing Laurentian-age rifting through Appalachian-Ouachita structures, Thomas (2006) interpreted that inheritance would be on a lithospheric scale. This notion that the mantle lithosphere would be susceptible to inherited structures, just as the crust would be, is in keeping with several studies highlighting the complete lithosphere as playing a part in deformation (e.g., Vauchez et al., 1997, 1998; Holdsworth et al., 2001; Bendick and Flesch, 2013; Li et al., 2016). In studying why continents seem to break-up parallel to orogenic belts, Vauchez et al. (1997) proposed that a pervasive fabric exists in the mantle lithosphere from ancient collisional events that can guide the propagation of continental rifts. Although the mantle lithosphere has been inferred to control rifting within the Wilson Cycle, the region has not had the same attention as the crust in terms of the evolution of the lithosphere. This is due, in part, to the difficult nature of studying the mantle lithosphere through imaging methods. However, recent advances have seen a substantial increase in research into the sub-crustal lithosphere.

Imaging the mantle lithosphere

Afonso et al. (2016) described the range of approaches used to study the lithosphere and upper mantle: teleseismic tomography (e.g., see Evans and Achauer (1993), Granet et al. (1995), Rawlinson et al. (2006)); surface-wave tomography (e.g., see Pasyanos and Nyblade (2007), Yang et al. (2008), Fishwick et al. (2008), Agius and Lebedev (2013)); gravity modelling (e.g., see Zeyen and Fernàndez (1994), Torne et al. (2000), Ebbing et al. (2006), Chapell and Kusznir (2008), Tašárová et al. (2009)); electromagnetic methods (e.g., see Heinson (1999), Jones (1999), Jones et al. (2009), Evans et al. (2005), Evans et al. (2011), and Meqbel et al. (2014)); local earthquake tomography (e.g., Aki and Lee (1976), Eberhart-Phillips (1990), and Kissling et al. (1994)); and receiver function studies (e.g., Yuan et al. (2006), Kawakatsu et al. (2009), Rychert and Shearer (2011), Kind et al. (2012)).

The increase in the number of high-resolution large-scale seismic arrays used in studies across the world has allowed for a clearer image of the deep lithosphere. The successful Lithoprobe project lasted from 1984 to 2005 and produced over 1500 publications on the evolution of the northern North American lithosphere. EarthScope initiated a 15-year programme of USArray, which consisted of the deployment of temporary and permanent seismic stations across the United States (comprising a Transportable Array, a Flexible Array, a (permanent) reference network and a magnetotelluric facility). The dense, moving network allowed for an unprecedented increase of image resolution of the North American lithosphere (e.g., Schaeffer and Lebedev, 2014). Other recent high resolution networks include (but are by no means limited to): the AFRICA Array (e.g., O’Donnell et al. 2016); the WOMBAT seismic array (e.g., Rawlinson and Fishwick, 2011); the M.A.G.I.C. array studying the crust and upper mantle of the Appalachian mountains; the ocean-based MERMAID project (Mobile Earthquake Recorder in Marine Areas by Independent Divers) uses floating receivers to image the deep earth (e.g., Hello et al., 2011); DANA (Dense Array in Northern Anatolia), imaging northern Turkey tectonics (e.g., Kahraman et al., 2015); the POLARIS (Portable Observatories for Lithospheric Analysis and Research Investigating Seismicity) array in Canada (e.g., Bastow et al., 2013); and the China National Digital Seismic Network (CNDSN) (e.g., Niu and Li, 2011; Bao et al., 2013).

This increase in research using large-scale imaging studies, alongside new techniques in acquisition and data processing (cf. Romanowicz, 2003; Artemieva et al., 2006; Rawlinson et al., 2010; Liu and Gu, 2012; Kuvshinov and Semenov, 2012) has also allowed structures below the Moho to be seen, with a multi-observable approach often built into the studies permitting corroboration of findings (e.g., deploying seismic and magnetotelluric stations). Results from new post-processing techniques of receiver function data have been encouraging (e.g., Rasendra et al., 2014; Tauzin et al., 2016; Park and Levin, 2016a; 2016b). The combination of receiver function and shear-wave splitting analysis on dense cross-fault arrays, as described in Rasendra et al. (2014), has been able to better characterize and understand the mechanics of large-scale strike-slip faults from the surface to the bottom of the lithosphere. When there is high-resolution imaging below the Moho, heterogeneities in the mantle lithosphere are ubiquitous (e.g., Rawlinson and Fishwick, 2011; Bastow et al., 2013; Schiffer et al., 2014, 2015, 2016; Schaeffer and Lebedev, 2014; Rasendra et al., 2014; Bao et al., 2014; Kahraman et al., 2015; Hopper and Fischer, 2015; Tauzin et al., 2016; Park and Levin, 2016a, 2016b; Biryol et al., 2016; Boyce et al., 2016; Dave et al., 2016). The relevance of these structures is currently being debated, but ultimately an understanding of them will help determine the role of the mantle lithosphere in the theory of plate tectonics.

Unravelling the tectonic impact of the mantle lithosphere

Through seismic imaging and geochemical analysis, the mantle lithosphere has been known to be disturbed or “scarred” for many years (e.g., Wendlandt et al., 1993; Lee et al., 2001; Yuan and Romanowicz, 2010; Lee et al., 2011), with deep inherited structures often interpreted to be the result of closure of ocean basins and continental collisions (e.g., Flack and Warner, 1990; Klemperer and Hobbs, 1991; Lie and Husebye, 1994; Morgan et al., 1994; Guellec et al., 1990; Pfiffner, 1992; Calvert et al., 1995; Calvert and Ludden, 1999; Cook et al., 1999; van der Velden and Cook, 2002; Cook, 2002; Cook and Vasudevan, 2003; White et al., 2003; Cook et al., 2004; van der Velden and Cook, 2005; Schiffer et al., 2014, 2015, 2016). The ages of these mantle lithosphere damage structures vary, with some features (Figure 3) thought to be of Archaean age (e.g., Calvert et al., 1995).

Although subduction scars have often been highlighted as a reason for the seismic visualization of mantle lithosphere reflectivity (e.g., Calvert et al., 1995; van der Velden and Cook, 2002; Cook, 2002), other processes exist that could create structures within the lithosphere. Van der Velden and Cook (2005) outline a number of other possibilities, including: mafic intrusions into the mantle (Steer et al., 1998); shear zones (Smythe et al., 1982; Warner and McGeary, 1987; Reston, 1990; McBride et al., 1995; Abramovitz et al., 1998); relict crustal fabrics and/or Moho

(Snyder, 1990; Cook and Vasudevan, 2003); and the lithosphere-asthenosphere boundary (Steer et al., 1998b).

The propensity of continents to break apart parallel to ancient orogenic belts also indicates a role of inherited structures in controlling tectonics, with rheological heterogeneity and mechanical anisotropy playing a factor (Vauchez et al., 1997, 1998). Furthermore, plate tectonic processes such as extensional stresses and plate bending prior to subduction have been suggested to weaken the rheology of oceanic lithosphere through the percolation of low-degree melts in metasomatic processes (Pilet et al., 2016). Taking such discussions into consideration, it is appropriate to interpret the seismic imaging of scarring to be regions of weakness in the continental mantle (e.g., Linckens et al., 2015; Heron et al., 2016a).

The role of grain damage in tectonic processes is also a method by which weakening could occur in the mantle lithosphere. In recent studies, Heron et al. (2016a, 2016b) interpret the seismic imaging of mantle lithosphere heterogeneities to be ancient deformation, with the reduction in grain size acting as a weak plane (Bercovici and Ricard, 2014). Lithospheric damage related to inheritance has been inferred to remain weak over very long timescales (Audet and Bürgmann, 2011), allowing ancient processes related to Archean scarring to be considered in present-day tectonics. At present, further constraints from the geological history of a region are required to unravel the processes related to the generation of mantle lithosphere heterogeneities and their impact on crustal tectonics. Numerical modelling has been shown to be useful in adding to the discussion on this topic of mantle lithosphere processes, an example of which (Heron et al., 2016b) is discussed below. Heron et al. (2016b) presented 2-D numerical experiments of continental convergence to generate intraplate deformation from inherited lithospheric structures (Figure 4a), exploring the limits of continental rheology to understand the dominant lithosphere layer across a broad range of geological settings.

Constraints from numerical modelling

The numerical experiments in Heron et al. (2016b), with some results shown here in Figure 4, were modelled using the two-dimensional, thermal-mechanical finite element numerical code SOPALE (Fullsack, 1995), which implements an Arbitrary Lagrangian-Eulerian (ALE) method to solve for the deformation of high Prandtl number incompressible viscous-plastic media. The models consider convergence in a stable (i.e., strong) (Burov and Watts, 2006) continental crust and mantle lithosphere setting (e.g., jelly sandwich rheology, Figure 1a) where the majority of mantle lithosphere scars are found (e.g., Steer et al., 1998a; Heron et al., 2016a). The model setup allows for a heterogeneous lithosphere, with a number of different weak zones in both the crust and mantle lithosphere (Figure 4a).

In Figures 4b–4e, crustal and mantle lithosphere inheritance is prescribed from Figure 4a as shown by the white scars and red heterogeneity, respectively. This configuration of the upper crust and lower crust weak zones permits easy identification of which layer is controlling deformation. After considerable shortening (in keeping with the extent of similar tectonic scenarios) (e.g., Cowgill et al., 2003), crustal thickening and faulting, key characteristics of intraplate orogenesis, are shown in models that feature upper crust (UC) or lower crust (LC) scars (Figures 4b and 4c). The implementation of a weak scar in the mantle lithosphere (overlain by a heterogeneous crust) dominates tectonics for this jelly sandwich rheology (Figure 4d). The models suggest that the impact of crustal scars is minimal when in the presence of a mantle lithosphere (ML) scar, as shown by comparing Figure 4d, featuring UC, LC, and ML scars, with Figure 4e, one ML scar only.

By implementing a 'crème brûlée' rheology (e.g., Figure 1b), featuring a weak mantle lithosphere and strong crust, it is found that heterogeneities within the mantle lithosphere become ineffective in controlling tectonics (Figure 4f). We posit that if the continental mantle is the strongest layer within the lithosphere, then such inheritance may have important implications for the development of tectonic processes in the Wilson Cycle (e.g., Holdsworth et al., 2001). Indeed, the rheological strength of the lithosphere may be imperative in analysing the cause and effect of large-scale tectonics (especially as scarring in the lithosphere is seen as ubiquitous). Furthermore, the models of Heron et al. (2016b) show that deformation driven by mantle lithosphere scarring can produce tectonic patterns related to intraplate orogenesis originating from crustal sources, making it difficult to unravel the cause of tectonic evolution while highlighting the need for a more formal discussion of the role of the mantle lithosphere in plate tectonics.

The Altyn Tagh Fault (ATF) in China illustrates the difficulty in unravelling tectonic cause and effect within the lithosphere. The tectonic history of China provides one reference to understand plate tectonics beyond plate boundaries with regards to the studies of Heron et al. (2016a, 2016b). Although there are many regions across the world where continents are subject to Wilson Cycle processes such as the continent accretion by closure of paleo-oceans between micro-plates, China is a unique reference as the far-field convergent stress from the Indian–Eurasian collision is relatively recent and ongoing (Figure 5a). The Altyn Tagh Fault (ATF), on the northern margin of the Tibetan Plateau, has a distinct present-day ML heterogeneity linked to a continent–continent suture (Cowgill et al., 2003). The ATF accommodates some of the convergence between the Indian and Eurasian plates (Zhang et al., 2014) and is characterized by localized deformation that has produced ~475 ± 70 km of staggered displacement since the mid-Oligocene (Cowgill et al., 2003). Although focal mechanisms of earthquakes close to the ATF show strike–slip motion, compressional processes account for earthquakes to the south (Zhang et al., 2014), with numerous thrust faults also inhabiting the area (Figure 5b). Geophysical studies of the ATF show deformation that penetrates the entire crust to link to heterogeneous structures in the ML (Wittlinger et al., 1998; Zhao et al., 2006; Zhang et al., 2014) (Figure 5c).

Could the ATF be interpreted as a ML scar originating as a continent–continent collision in the Palaeozoic (Sobel and Arnaud, 1999) that controls intraplate deformation during periods of compression (with the most recent episode starting in the Oligocene resulting from the India–Eurasia collision)? Or is it that the ML scar is a result of crustal deformation impinging on the deeper lithosphere? The ability of deep lithospheric heterogeneous structures to exist over long periods in stable continental settings allows for a new mechanism for intraplate evolution (following external forcing). If, as an example, the ATF has a long-lasting ML scar from a continental collision that is controlling the crustal evolution, then plate tectonics may indeed display timeless (‘perennial’) processes (e.g., Heron et al., 2016a) with plate boundaries never really disappearing. As such, an increase in intraplate orogenesis would be observed during future (and past) periods of global compression and extension (that is, supercontinent formation and dispersal).

However, deep inheritance as a source of intraplate deformation (and as a process within the Wilson Cycle as a whole) is not a closed subject. One reason for this is the ambiguity in the rheological properties of the scars “frozen” into the lithosphere. Schiffer et al. (2016) interpret mantle lithosphere scarring on the continental margin of East Greenland to be of higher density than the surrounding mantle material, with Petersen and Schiffer (2016) providing modelling on the topic. However, a number of studies have discussed the weakening impact of tectonic processes on the lithosphere to facilitate continental rifting (Dunbar and Sawyer, 1988, 1989). Furthermore, the subduction of crustal material into the mantle through ancient processes could increase volatiles to the lower lithosphere, weakening the seismically imaged scarred material (Pollack, 1986).

Aside from numerical modelling, the wider discussion on what we can ‘see’ in the mantle lithosphere and what we can infer from structures has been bolstered by a great number of seismic studies in recent years.

Constraints from seismic studies

Figure 6a shows examples of regions where mantle lithosphere heterogeneities (yellow circles) have been inferred, compiled from a previous map by Steer at al. (1998a) and updated to include more recent studies (e.g., Cook et al., 1999; van der Velden and Cook, 2005; Yang et al., 2003;

Hopper and Fischer, 2015; Kahraman et al., 2015; Schiffer et al., 2016). As discussed, the increase in high resolution imaging studies has increased the discovery of such structures in recent years. For an interpretation of the 2D geometry of the heterogeneities, Figure 6b gives an estimation of diagonal length of a mantle lithosphere scar (from a 2D horizontal and vertical component), with accompanying angle from the horizontal, for eight examples of mantle lithosphere heterogeneities (from Heron et al., 2016b). Below we outline a number of studies indicating an increased 'visibility' into the mantle lithosphere.

For example, the high-density seismometer array on the North Anatolian fault (NADA) showed horizontal structural variations in the crust and upper mantle on scales of 10 km and 20 km, respectively (Kahraman et al., 2015). Using USArray data, Hopper and Fischer (2015) applied converted wave imaging to the northern US craton to reveal mid-lithospheric discontinuities within the thick, high-velocity mantle. Their findings show that volatile rich layers could become 'frozen into' the mantle lithosphere as the lithosphere cools.

A clear link between plate tectonics, inheritance, and intraplate tectonics has been highlighted in Biryol et al. (2016), which presents new tomographic images of the south-eastern United States, revealing large-scale structural variations in the upper mantle. The origin of these structures is inferred to be a product of earlier episodes of continental collision and breakup, suggesting that the Wilson Cycle can generate long-lasting features within the mantle. Biryol et al. (2016) also discuss that plate strength and pre-existed inherited structures are important mechanisms that may be controlling ongoing tectonism in the region, as well as the multiple zones of seismicity.

The WOMBAT transportable seismic array in southeast Australia has imaged multiple lithospheric structures, as described in Rawlinson and Fishwick (2011). The mantle lithosphere is shown to have a wealth of features related to the geology and tectonic history of the region. The discovery of structures in certain areas related to lithospheric thinning, as well as Paleozoic provinces at depth in other regions, may have profound implications for the break-up of Australia and Antarctica. Furthermore, the use of new P and S wave tomography has been able to constrain upper mantle structures beneath southeast Canada and the northeast USA, a region spanning three quarters of Earth's geological history (Boyce et al., 2016). The ability to differentiate wave speeds within a medium to a finer degree has allowed for better understanding of how stable cratonic keels may have formed (Boyce et al., 2016), as new interpretations can be made on the processes that could cause lateral strength variations within the mantle lithosphere under North America (based on the tectonic history). It is the high-resolution illumination of the sub-crust (e.g., Rawlinson and Fishwick, 2011; Boyce et al., 2016) that can generate discussion on Wilson Cycle processes (continental break-up, craton stabilization) that were never possible in the past.

An abrupt seismic velocity wave speed transition in the mantle lithosphere from craton to Cordillera in western Canada was recently documented by Bao et al. (2014). This transition was interpreted to be related to the modification of the mantle lithosphere through Wilson Cycle dynamics, namely subduction zone interaction (Bao et al., 2014). Their discussion highlighted the possibility of small-scale convection initiated by a zone of weakness between the craton and the thickened lithospheric margin. Another recent important paper is the work of Dave et al. (2016), which presents a three-dimensional shear wave velocity model beneath the Wyoming craton constrained from Rayleigh wave data. Their model provides the first seismic evidence for complex small-scale mantle convection beneath the Wyoming craton, with a high-velocity anomaly having a dripping shape in central Wyoming extending to 200 - 250 km depth (indicating mantle downwelling and lithosphere erosion).

Chamberlain et al. (2014) studied the San Andreas Fault and analysed the strain history of the upper mantle. Through the comparison of the long-term finite strain field in the mantle and the surface strain-rate field, respectively inferred from fast polarization directions of seismic phases (SKS and SKKS) and GPS data, Chamberlain et al. (2014) inferred that the San Andreas Fault extends to depth, likely through the entire lithosphere, with the possibility of the asthenosphere and tectonic plate being coupled. Asthenosphere mantle flow generating dynamic topography through vertical motions has also been investigated as a cause of lithosphere tectonics. Becker at al. (2014) highlighted western US intermountain seismicity as being caused by changes in upper mantle flow. The study inferred that mantle flow plays a significant and quantifiable part in shaping topography, tectonics, and seismic hazard within intraplate settings. If intraplate tectonics can be added into the Wilson Cycle dynamics, as we consider is sensible (e.g., Heron et al., 2016b), then the influence of the mantle lithosphere and convecting mantle on long-term and short-term tectonics is an important factor that is becoming clearer in recent years.

Discussion and Conclusions

In this review, we have outlined the current research on the role of the mantle lithosphere in causing tectonic deformation, alongside highlighting the potential of the deep lithosphere in infiltrating every aspect of plate tectonics processes. As such an endeavour often leaves more questions than answers, we have compiled open questions on the role of the mantle lithosphere in the Wilson Cycle:

- How pervasive is localized deformation within the mantle lithosphere? For example, are deeps scars abundant, but just not imaged; or is the imaging fairly accurate in showing lithosphere that is less scarred than the upper crust?

- Are the structures that are 'visible' in the continental mantle lithosphere of large-scale tectonic importance? Do they indicate zones of weakness (e.g., (Bercovici and Ricard, 2014) or strength (e.g., Schiffer et al., 2016)? Can they be treated as pathways of future plate tectonic deformation?

- Do all Wilson Cycle continent collision and break-up events generate major mantle lithosphere scale structures (e.g., Biryol et al., 2016)?

- How can we differentiate among the causes of lithosphere scale deformation? For example, can we differentiate between mantle lithosphere structures caused by deformation originating in the crust and crustal deformation caused by reactivating mantle lithosphere structures?

- What is the role of isolated mantle volatiles being ‘frozen’ into the mantle lithosphere (e.g., Hopper and Fischer, 2015)? Are non-continuous zones of volatiles widespread across the whole of continental mantle lithosphere or simply localized features?

- Is the large-scale rheological layering of the lithosphere more important in permitting the initiation of tectonic deformation than features within the lithosphere (e.g., scarring and inherited structures)? Or is it that lithosphere rheology and small features must be considered as a coupled system (e.g., Heron et al., 2016b)?

At the centre of these questions is the rheological make-up of the mantle lithosphere and the layering of the lithosphere as a whole (as discussed in the introductory section). Future work is required to constrain the strength layering within the continental lithosphere, and to what spatial extent such an environment can be applied.

The introduction of intraplate deformation to the Wilson Cycle is something that we put forth here and in a previous manuscript (Heron et al., 2016b). We would argue that the Wilson Cycle should be expanded to include intracontinental tectonics. Furthermore, we would highlight the notion that plate boundaries may never truly disappear through inherited structures. A tenet of the conventional theory of plate tectonics (and indeed the Wilson Cycle) is that crustal deformation is confined to near the boundaries of plates. Recent work on inheritance implies that this remains true for general planetary deformation as ML scars (that can control tectonic evolution) in a continent interior may originate from ancient plate boundary deformation (e.g., Heron et al., 2016a). In this way, ancient and present-day plate boundaries could be represented together as latent and active boundaries. A global map of perennial plate tectonics (Figure 6) presents a redefined illustration of tectonic activity and modifies the conventional theory of plate tectonics (in keeping with the recent findings of Vauchez et al., (1997), Rawlinson and Fishwick (2011), Bercovici and Ricard (2014), Leng and Gurnis (2015), Dave et al. (2016), Boyce et al. (2016)).

Although images of the sub-crustal lithosphere are becoming more commonplace, there are areas where such studies are not possible due to accessibility and expense. An interesting alternative is the work of Flesch and Bendick (2012) who consider the relationship between surface kinematics and deformation of the whole lithosphere. Flesch and Bendick (2012) used 3-D numerical models to find a relationship between tectonics at the surface and deformation throughout the crust and mantle lithosphere, through changing the lithosphere strength profile (e.g., Figure 1). Their study found that where viscosity is both discontinuous and differs by much more than an order of magnitude between the upper crust and mantle lithosphere, information about both force balance and rheology are absent from the surface deformation. It is therefore difficult to estimate either the dynamic or mechanical state of the lithosphere through surface observations (Flesch and Bendick, 2012).

The use of numerical modelling will help to understand further the complex nature of mantle lithosphere scarring, and this, as well as the interaction with the crust above, may be better understood in three dimensions (e.g., Chen and Gerya, 2016). Numerical modelling of a lithosphere with a ‘lasting memory’, following on from the work of Bercovici and Ricard (2014) (and others), will become more commonplace in plate tectonic studies in order to meet the requirement of inherited structures. If inherited structures are to evolve and dictate lithosphere evolution, then numerical models will need to model long timescales to take into consideration past dynamics in order to understand present and future evolution (e.g., Bercovici and Ricard, 2014).

As the imaging of the lithosphere becomes clearer, the assumed strength profile of tectonic plates is becoming more complex (e.g., Figure 1). At the same time, the inherent strength of the structures within the mantle lithosphere is not well known. Work is required to fully understand the nature of the mantle lithosphere heterogeneities, as mantle lithosphere scarring has been interpreted to be either areas of weakness (e.g., Dunbar and Sawyer, 1988, 1989; Pollack, 1986; Bercovici and Ricard, 2014; Linckens et al., 2015; Heron et al., 2016) or strength (e.g., Schiffer et al., 2016; Boyce et al., 2016), which may alter the deformation evolution (e.g., Heron et al., 2015b). The integration of mantle geochemistry into studies of lithosphere deformation will be important in this discussion, in particular the evolution of grain damage over time (e.g., Bercovici and Ricard, 2014). The link between grain-damage hysteresis and plate tectonic states may allow for a new analysis on how our planet may evolve differently to other terrestrial bodies (Bercovici and Ricard, 2016).

As body of evidence grows for the importance of the mantle lithosphere in plate tectonic processes (e.g., Vauchez et al., 1997; Holdsworth et al., 2001; Rawlinson and Fishwick, 2011; Bercovici and Ricard, 2014; Leng and Gurnis, 2015; Dave et al., 2016; Boyce et al., 2016; Heron et al., 2016a), it would be prudent for future work to consider the global and/or local aspect of their discoveries. The interpretation of the role of the mantle lithosphere should be considered as such: is the fundamental rheological composition of the mantle lithosphere important on a global scale, or does the evolution of the lithosphere in a given area present specific examples of mantle lithosphere importance? This distinction between a globally applicable discovery and local evolution may be important in the analysis of the role of the mantle lithosphere in the Wilson Cycle.

The Wilson Cycle (Figure 2) describes the closure and opening of oceanic basins (e.g., Wilson, 1966; Dewey and Burke, 1974), where continental margins are deformed and weakened over time. The geological and geophysical mechanisms within the Wilson Cycle encapsulate our conventional theory of plate tectonics, with structural inheritance in the tectonic plates playing a strong role in the evolution of the lithosphere (e.g., Holdsworth et al., 2001). Heron et al. (2016a) argue that if intraplate deformation can be linked to inherited structures from ancient plate tectonic events, then deformation within continental margins should also be part of a wider Wilson Cycle (Figure 2). Furthermore, the role of the mantle lithosphere as a source of pre-existing structures that could influence tectonics is coming to the forefront of tectonic dynamics (e.g., Vauchez et al., 1997; Holdsworth et al., 2001; Rawlinson and Fishwick, 2011; Bercovici and Ricard, 2014; Leng and Gurnis, 2015; Dave et al., 2016; Boyce et al., 2016; Heron et al., 2016a), as well the role of the deep lithosphere (and sub-lithosphere mantle) in surface tectonics (e.g., Chamberlain et al., 2014; Becker et al., 2015; VanderBeek et al., 2016). High-resolution seismic imaging surveys over the past decade has found heterogeneous structures within the mantle lithosphere to be somewhat ubiquitous (e.g., Rawlinson and Fishwick, 2011; Bastow et al., 2013; Schiffer et al., 2014, 2015, 2016; Schaeffer and Lebedev, 2014; Rasendra et al., 2014;

Bao et al., 2014; Kahraman et al., 2015; Hopper and Fischer, 2015; Tauzin et al., 2016; Park and Levin, 2016a, 2016b; Biryol et al., 2016; Boyce et al., 2016; Dave et al., 2016). There is a strong case for the importance of the mantle lithosphere in Wilson Cycle processes, through inherited structures, with an incentive to look deeper at how tectonic plates evolve.

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FIGURE CAPTIONS

Figure 1. Schematic view of alternative first-order models of strength through continental lithosphere (from Bürgmann and Dresen, 2008). In the upper crust, frictional strength increases with pressure and depth. In the two left panels a coefficient of friction following Byerlee's law and hydrostatic fluid pressure (ratio of pore pressure to lithostatic pressure $\lambda = 0.4$) are assumed in a strike-slip tectonic regime. In the right panel, low friction due to high pore fluid pressure ($\lambda = 0.9$) is assumed. (a) A jelly sandwich strength envelope is characterized by a weak mid-to-lower crust and a strong mantle composed dominantly of dry olivine (Hirth and Kohlstedt, 2003). (b) The crème brûlée model posits that the mantle is weak (in the case shown resulting from a higher geotherm, adding water would produce a dramatic further strength reduction). The dry and brittle crust defines the strength of the lithosphere. (c) The banana split model considers the weakness of major crustal fault zones throughout the thickness of the lithosphere, caused by various strain weakening and feedback processes. Owing to small grain size in shear zones, deformation in the lower crust and upper mantle is assumed to be accommodated by linear diffusion creep (grain size of 50 μm).

Figure 2. The Wilson Cycle with the additional tectonic feature of intraplate deformation. Rifting (B), continental collision (D), and/or intraplate deformation (i) can leave lasting impressions on the crust and mantle. The importance of inherited crustal and mantle structures in influencing the tectonic pathway of deformation is shown by purple arrows. The figure shows that it is difficult to unravel the cause and effect on the lithosphere of Wilson Cycle processes. The references for the established pathway tectonic influence are as follows: [1] e.g., Holdsworth

et al. (2001); Holdsworth (2004); Thomas (2006); [2] e.g., Royden and Keen (1980), Davis and Kusznir (2004), Buiter et al. (2009), and Péron-Pinvidic et al. (2013); [3] e.g., Vauchez et al. (1997); [4] e.g., Flack and Warner (1990), Morgan et al. (1994), Lie and Husebye (1994), Calvert et al. (1995), Calvert and Ludden (1999), Ghazian and Buiter (2013), and Schiffer et al. (2014, 2016); [5] e.g., Tapponnier and Molnar (1975); [6] e.g., Dèzes et al. (2004), Avouac et al. (1993), Cowgill et al. (2003), Tapponnier and Molnar (1975), and Kahraman et al. (2015); [7] e.g., Stephenson et al. (2009); [8] e.g., Heron et al. (2016a). This figure is modified from Heron et al. (2016b).

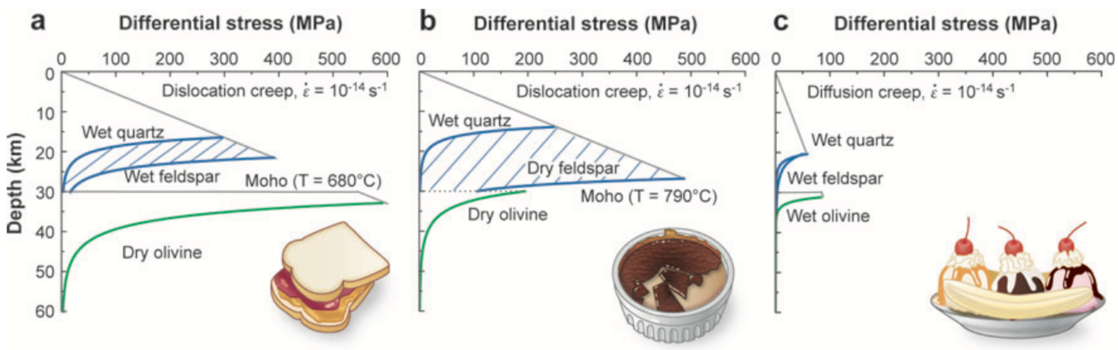
Figure 3. An example of a mantle reflection from Calvert et al. (1995). Line migration results of the Abitibi-Opatica survery (a) with interpreted results (b). The most prominent feature of the data is the band of mantle reflections that dip in the north to northwest direction beneath the Opatica belt. The mantle reflections intersect the Moho beneath the Abitibi-Opatica boundary mapped at the surface (Calvert et al., 1995).

Figure 4. Overview of numerical modelling results into continental intraplate deformation related to far-field compression in the presence of upper crust (UC), lower crust (LC), and mantle lithosphere (ML) heterogeneities. The full numerical simulation is performed with SOPALE across 600 km depth and 1500 km across. Rheological parameters are given in Heron et al. (2016b), with compression applied at 1 cm/yr. (a) Positions of scars used in the numerical study of Heron et al. (2016b). The scar length and angle are given in Figure 6b. The weak zones (scars) in the UC and LC (as shown in white) and ML (red). Panels (b) – (e) show deformation patterns related to a ‘jelly sandwich’ rheology similar to that of Figure 1a. Material deformation

(top) and visualization of the second invariant of the deviatoric strain rate tensor (bottom) after shortening for (b) model with UC scar only, (c) model with LC scar only, (d) model with all scars, and (e) model with a ML scar only. Top 100 km of the models are shown in a 3X vertical exaggeration. Models show that heterogeneities within the mantle lithosphere can control tectonics over shallower features in strong mantle lithosphere settings. Panel (f) shows the deformation of a continental interior for a crème brûlée (CB) lithosphere strength profile (generated through a hot Moho temperature). (f) shows the mantle lithosphere scar playing no role in deformation, highlighting the importance of lithosphere strength in tectonic evolution (e.g., Figure 1).

Figure 5. The suture zones of Chinese tectonics and the Altyn Tagh Fault (ATF) (from Heron et al. (2016a). (a) A topographic map of the different tectonic blocks with paleo-suture zones (white lines) of the India–Eurasia collision zone (suture zones from Watson et al., 1987). CAOB, Central Asia Orogenic Belt; L, Lhasa block; Q, Qaidam Basin; QI, Qiantang block; SQ, Songpan–Ganzi complex; TB, Tarim Basin. (b) Grey boxed region in (a) showing the ATF with strike-slip faulting denoted in black, with thrust faulting in white (Cowgill et al., 2003). NAF, North Altyn Fault. (c) Schematic seismic model of ATF (Wittlinger et al., 1998) from Zhang et al. (2015). Red and green regions indicate the crust and mantle, respectively. Regions that are more yellow or red in the model are low-velocity zones. Seismic line A to A0 is marked on b. This region may represent an instance of a mantle lithosphere heterogeneity controlling intraplate crustal deformation through far-field compressional forcing (e.g., Heron et al., 2016a).

Figure 6. (a) A perennial plate tectonic map showing examples of regions where mantle lithosphere heterogeneities (yellow circles) have been inferred, compiled from a previous map by Steer at al. (1998a) and more recent studies (Cook et al., 1999; van der Velden and Cook, 2005; Yang et al., 2003; Hopper and Fischer, 2015; Kahraman et al., 2015; Schiffer et al., 2016), alongside some possible paleo-plate boundary locations (yellow lines) (as modified from Holt et al., 2015). (b) Estimation of mantle lithosphere scar length and angle from horizontal for eight examples of mantle lithosphere heterogeneities (from Heron et al., 2016b).



a
Differential stress (MPa)
0
100
200
300
400
500
600
Depth (km)
10
20
30
40
50
60
Dislocation creep, $\dot{\varepsilon} = 10^{-14}$ s^{-1}
Wet quartz
Wet feldspar
Moho (T = 680°C)
Dry olivine
b
Differential stress (MPa)
Dislocation creep, $\dot{\varepsilon} = 10^{-14}$ s^{-1}
Wet quartz
Dry feldspar
Moho (T = 790°C)
Dry olivine
c
Differential stress (MPa)
Diffusion creep, $\dot{\varepsilon} = 10^{-14}$ s^{-1}
Wet quartz
Wet feldspar
Wet olivine

Figure 1.

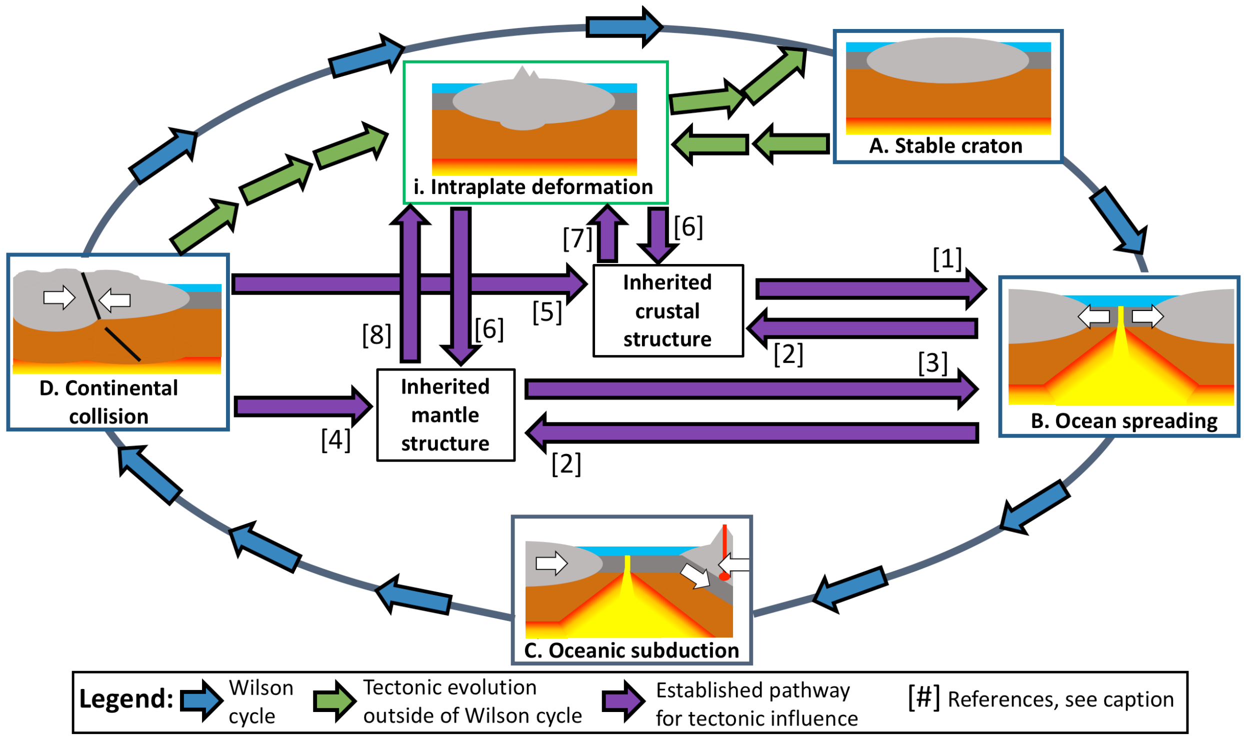


Figure 2.

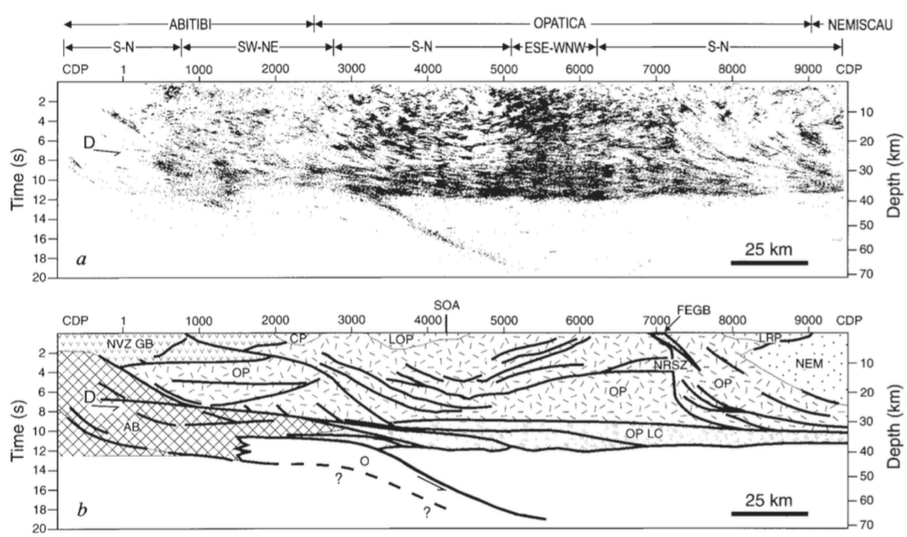
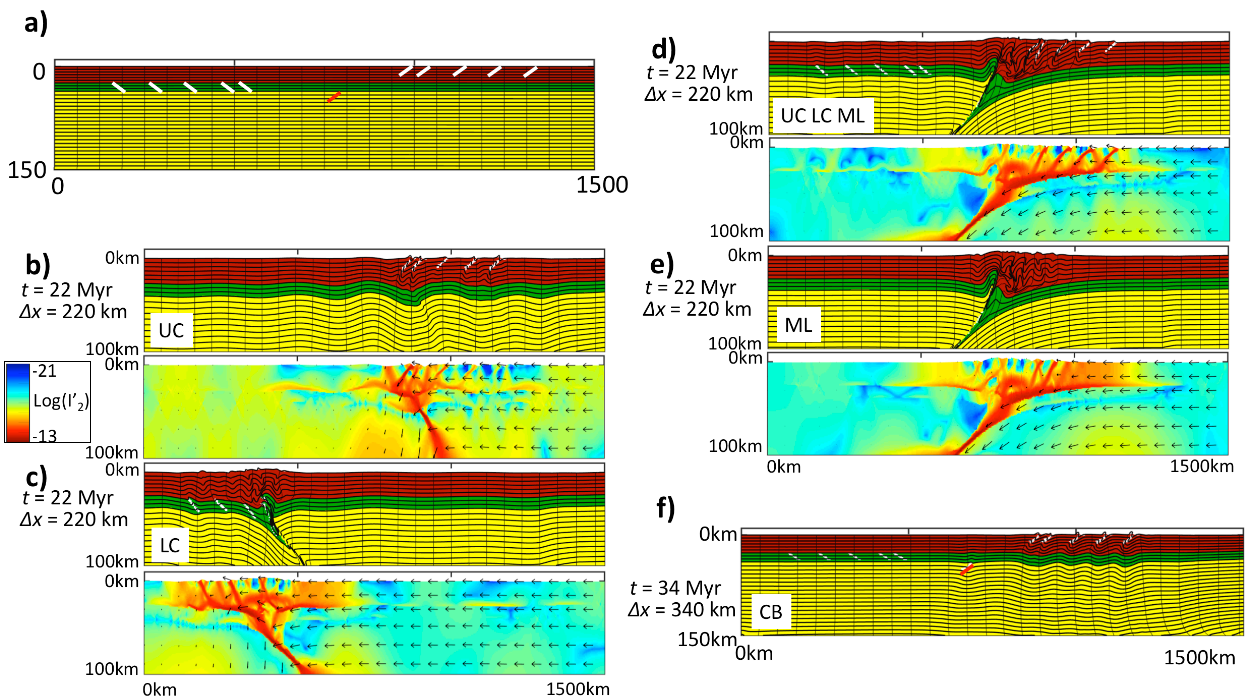


Figure 3.



a)
0
150
0
1500
b)
t = 22 Myr
Δx = 220 km
UC
0km
100km
-21
Log(I'2)
-13
c)
t = 22 Myr
Δx = 220 km
LC
0km
1500km
d)
t = 22 Myr
Δx = 220 km
UC LC ML
e)
t = 22 Myr
Δx = 220 km
ML
f)
t = 34 Myr
Δx = 340 km
CB
150km

Figure 4.

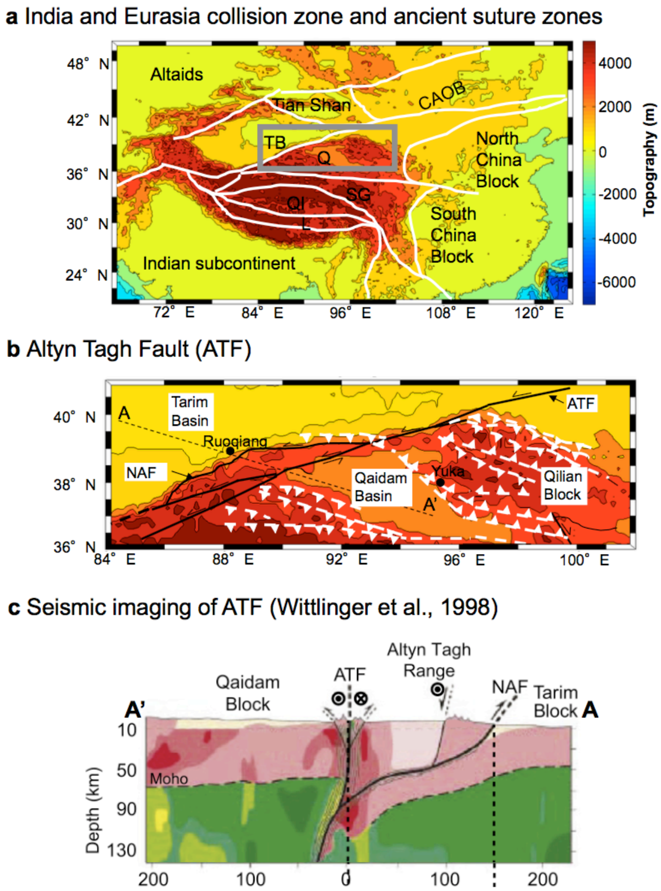

a India and Eurasia collision zone and ancient suture zones
Altaids
Tian Shan
CAOB
TB
Q
North China Block
SG
QI
L
South China Block
Indian subcontinent
48° N
42° N
36° N
30° N
24° N
72° E
84° E
96° E
108° E
120° E
4000
2000
0
-2000
-4000
-6000
Topography (m)
b Altyn Tagh Fault (ATF)
Tarim Basin
A
Ruoqiang
NAF
Qaidam Basin
Yuka
A'
Qilian Block
ATF
40° N
38° N
36° N
84° E
88° E
92° E
96° E
100° E
c Seismic imaging of ATF (Wittlinger et al., 1998)
Altyn Tagh Range
Qaidam Block
ATF
NAF
Tarim Block
A'
A
Moho
Depth (km)
10
50
90
130
200
100
0
100
200

Figure 5.

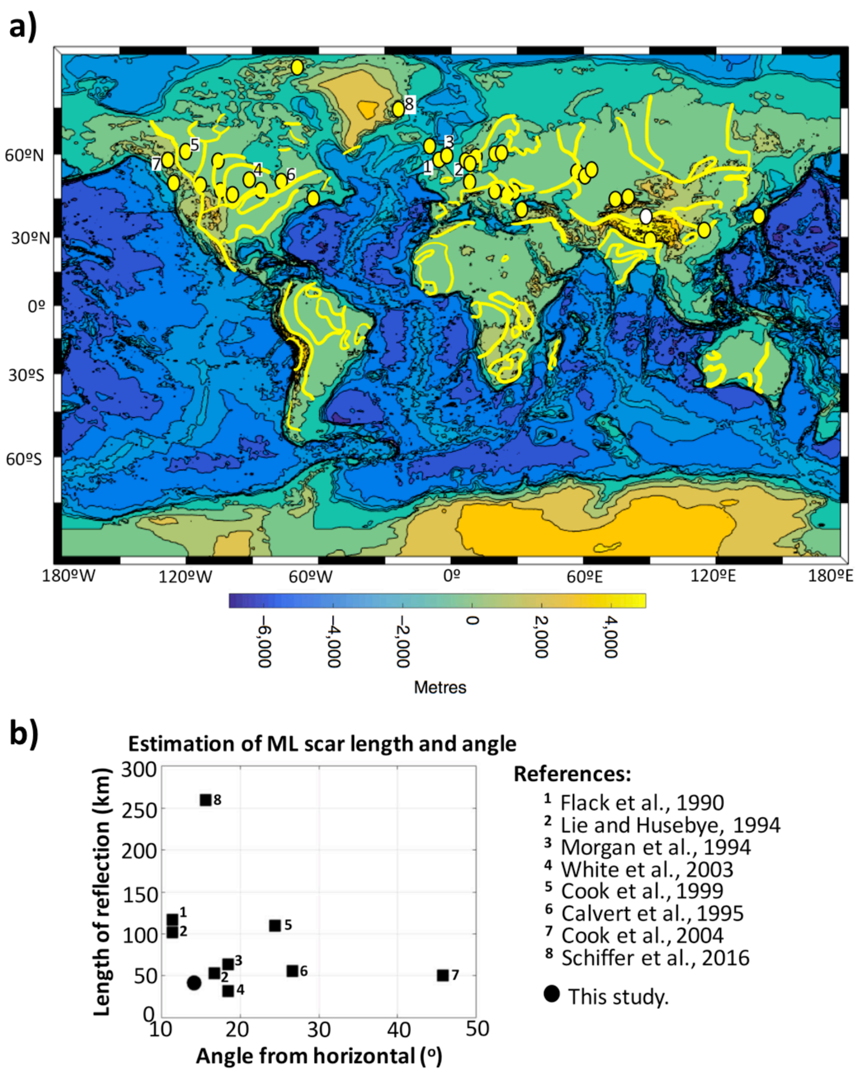

a)
60ºN
30ºN
0º
30ºS
60ºS
180ºW
120ºW
60ºW
0º
60ºE
120ºE
180ºE
-6,000
-4,000
-2,000
0
2,000
4,000
Metres
b)
Estimation of ML scar length and angle
Length of reflection (km)
Angle from horizontal (º)
References:
1 Flack et al., 1990
2 Lie and Husebye, 1994
3 Morgan et al., 1994
4 White et al., 2003
5 Cook et al., 1999
6 Calvert et al., 1995
7 Cook et al., 2004
8 Schiffer et al., 2016
This study.

Figure 6.